A Numerical Simulation on Japan/East Sea (JES) Thermohaline Structure and Circulation

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ABSTRACT

The seasonal circulation and thermohaline structure in the Japan/East Sea (JES) were studied numerically using the Princeton Ocean Model (POM) with horizontal resolution $(1/12)^\circ \times (1/12)^\circ$ and 23 sigma levels conforming to a relatively realistic bottom topography. A two-step initialization technique is used. During the first step (restoring run), the POM is integrated for two years from zero velocity and January climatological temperature and salinity fields with climatological monthly mean surface wind stress. The final states of the restoring run are taken as initial conditions for the second step (simulation run). During the simulation run, the POM is integrated again for two more years with climatological monthly mean surface wind stress, net heat flux, and fresh-water flux from the COADS data. The simulated temperature, salinity, and velocity fields are consistent with observational studies reported in Part 1 of this paper.

1. INTRODUCTION

The Japan Sea, known as the East Sea in Korea, has a steep bottom topography (Fig. 1) that makes it a unique semi-enclosed ocean basin overlaid by a pronounced monsoon surface wind. The Japan/East Sea, hereafter referred to as JES, covers an area of 10^6 km^2. It has a maximum depth in excess of 3,700 m, and is isolated from open oceans except for small (narrow and shallow) straits, which connect the JES with the North Pacific through the Tsushima/Korean and Tsugaru Straits and with the Okhotsk Sea through the Soya and Tatar Straits. The JES contains three major basins called the Japan Basin (JB), Ulleng/Tsushima Basin (UTB), and Yamato Basin (YB), and has a high central plateau called the Yamato Rise (YR). The JES has great scientific interest as a miniature prototype ocean. Its basin-wide circulation pattern, boundary currents, Subpolar Front (SPF), mesoscale eddy activities and deep water formation are similar to those in a large ocean.

The seasonal variability of the JES thermohaline structure has been studied by many investigators (Gong, 1968; Isoda and Saitoh, 1993; Isoda et al., 1991; Isoda, 1994; Kano, 1980; Maizuru Mar. Observ., 1997) using limited data sets. For example, Isoda and Saitoh (1993) analyzed satellite infrared (IR) images in the western part of the JES and the routine hydrographic survey completed by the Korea Fisheries Research and Development Agency in 1987 for the winter and the spring. They found that a small meander of a thermal front originates from the Tsushima/Korea Strait near the Korean coast and gradually grows into an isolated warm eddy with a horizontal scale of 100 km. The warm eddy moves slowly northward from spring to summer.

Recently, Chu et al. (1998a; 1999a) reported the seasonal occurrence of JES eddies from the composite analysis on the U.S. National Centers for Environmental Prediction

(NCEP) monthly SST fields (1981-1994). For example, they identified a warm center appearing in late spring in the East Korean Bay. Chu et al. (2001a, b) further reported the seasonal variation of the thermohaline structure and inverted circulation from the Navy's unclassified Generalized Digital Environmental Model (GDEM) temperature and salinity data on a 0.5° × 0.5° grid. The GDEM for the JES was built on historical 136,509 temperature and 52,572 salinity (1930-1997) profiles. A three-dimensional estimate of the absolute geostrophic velocity field was obtained from the GDEM temperature and salinity fields using the P-vector method. The climatological mean and seasonal variabilities of the thermohaline structure and the inverted currents such as the SPF, the mid-level (50 to 200 m) salty tongue, the Tsushima Warm Current (TWC) and its bifurcation were identified.

Numerical studies on the JES circulations started in the early 1980s. Various types of models were used such as the multi-layer model (Sekine, 1986, 1991; Kawabe, 1982; Yoon, 1982a,b; Seung and Nam, 1992; Seung and Kim, 1995), the Modular Ocean Model (MOM) (Kim and Yoon, 1994; Holloway et al., 1995; Kim, 1996; and Yoon, 1996; Yoshikawa et al., 1999), rigid-lid z-level model (Yoshikawa et al., 1999), the Miami Isopycnal Coordinate Model (MICOM) (Seung and Kim, 1993) and the Princeton Ocean Model (POM) (Mooers and Kang, 1995; Chu et al., 1999b). Most of the numerical efforts are concentrated on simulating the basin-wide circulation, the TWC bifurcation, and formation of the intermediate waters using idealized or restoring-type surface thermal forcing. There is lack of study on simulating seasonal variabilities of the JES circulation and thermohaline structure using flux-type thermal forcing.

In this study, the POM with flux-type surface forcing is used to simulate the seasonal variabilities of the JES circulation and thermohaline structure including the meandering SPF and eddies, the TWC bifurcation, the retroflection of the East Korea Warm Current (EKWC), the Liman Cold Current (LCC) and its penetration into the southwestern waters along the Korean coast. These features were identified in the first part of this paper using the Navy's GDEM data. The outline of this paper is as follows: A description of the JES current systems is given in Section 2. A depiction of the seasonal variation of atmospheric forcing is given in Section 3. The numerical ocean model and integration are depicted in Section 4. The simulated seasonal variability of temperature, salinity, and circulation are discussed in Sections 5, 6, and 7, respectively. In Section 8 we present our conclusions.

2. JES Current Systems

Most of the nearly homogeneous water in the deep part of the basin is called the Japan Sea Proper Water (Moriyasu, 1972) and is of low temperature and low salinity. Above the Proper Water, the TWC, dominating the surface layer, flows in from the East China Sea through the Tsushima/Korean Strait and carries warm water from the south. The LCC carries cold fresh surface water from the north and northeast (Seung and Kim, 1989; Holloway et al. 1995). The properties of this surface water are generally believed to be determined by the strong wintertime cooling coupled with fresh water input from the Amur River and the melting sea ice in Tatar Strait (Martin and Kawase, 1998). The LCC flows southward along the Russian coast, beginning at a latitude slightly north of Soya Strait, terminating off Vladivostok (Fig.2), and becoming the North Korean Cold Current (NKCC) after reaching the North Korean coast (Yoon, 1982).

The TWC separates into two branches which flow through the western and eastern channels of the Tsushima/Korea Strait (Kawabe, 1982a,b; Hase et al., 1999). The flow through the eastern channel closely follows the Japanese Coast and is called the Nearshore Branch (Yoon, 1982) or the first branch of TWC (FBTWC) (Hase et al., 1999). The flow through the western channel is called the EKWC, which closely follows the Korean coast until it separates near 37˚N into two sub-branches. The western sub-branch moves northward and forms a cyclonic eddy off the eastern Korean coast. The eastern sub-branch flows eastward to the western coast of Hokkaido Island, and becomes the second branch of the TWC (SBTWC).

The NKCC meets the EKWC at about 38˚N with some seasonal meridional migration. After separation from the coast, the NKCC and the EKWC converge and form a strong front that stretches in the zonal direction across the basin. The NKCC makes a cyclonic recirculation gyre in the north but most of the EKWC flows out through the Tsugaru and Soya Straits (Uda, 1934). The formation of NKCC and separation of EKWC are due to local forcing by wind and buoyancy flux (Seung, 1992). Large meanders develop along the front and are associated with warm and cool eddies.

Seung (1995) identified major features of the volume transport from earlier numerical modeling results. The transport pattern is largely determined by the upper layer circulation and characterized by a large-scale cyclonic recirculation gyre, in which the EKWC and the Nearshore Branch take part as the inflow-outflow system, and also includes the NKCC. A few hundred kilometers from the separation point, the EKWC forms an anticyclonic gyre. The gyre becomes stronger as the EKWC develops. On the other hand, the northern cyclonic gyre is very deep and is most significant in the winter when strengthened by the wind and buoyancy flux. It is usually called the JB gyre. The gyre, or the southward coastal current (NKCC) related to it, is deep enough to intrude southward beneath the EKWC most of the time. Seung also confirmed the summertime presence of a countercurrent beneath the Nearshore Branch.

3. Seasonal Variation of Atmospheric Forcing

3.1. General Description

The Asian monsoon strongly affects the thermal structure of the JES. During the winter monsoon season, a very cold northwest wind blows over the JES (Fig. 3a) as a result of the Siberian High Pressure System with a mean surface wind speed between 10 and 15 ms^{-1}. By late April, numerous frontally-generated events occur making late April and May highly variable in terms of wind speeds and amount of clouds. During this period storms originating in Mongolia may cause strong, warm westerlies (Fig. 3b). By late May and early June, a summer surface atmospheric low pressure system begins to form over Asia. Initially this low pressure system is centered north of the Yellow Sea producing westerly winds. In late June, this low begins to migrate to the west setting up the southwest monsoon that dominates the summer months. The winds remain variable through June until the Manchurian Low Pressure System strengthens. Despite the very active weather systems, the mean surface wind speed over the JES in summer (Fig. 3c) is between 3 and 4 m/s, which is weaker than in winter (Fig. 3a). By July, however, high pressure (the Bonin High) to the south and the low pressure over Manchuria produce southerly winds carrying warm, moist air over the East China Sea/Yellow Sea. In the summer, warm air and strong downward net radiation stabilize the upper layer of the JES and causes the surface mixed layer to shoal. October (Fig. 3d) is the beginning of the transition to winter conditions. The southerly winds weaken and the sea surface slope reestablishes its winter pattern.

Here, a climatological description of the surface net heat and fresh water fluxes over the JES is presented. The datasets used were the objectively analyzed fields of surface marine climatology and anomalies of fluxes of heat, momentum, and fresh water. The fields are derived from individual observations in the Comprehensive Ocean-Atmosphere Data Set (COADS) from 1945 to December 1989 and are analyzed on a 1° by 1° grid (da Silva et al., 1994).

3.2. Net Surface Heat Flux

Net surface heat flux is computed by

$$Q_{Net} = R_S - (R_L+Q_S+Q_L) \tag{1}$$

where R_S is the net downward shortwave radiation, R_L the net upward longwave radiation, Q_S the sensible heat flux, and Q_L the latent heat flux. Positive (negative) values of Q_{Net} indicate net heat gain (loss) of the ocean at the surface. The summer field is relatively homogeneous (140 to 160 W m^{-2}) throughout the JES, whereas a significant horizontal gradient increasing from the southeast (Japan coast) to the northwest (east Russian coast) exists for the rest of the seasons (Fig. 4). The ocean surface near the Tsushima/Korea Strait has the maximum heat loss of 400 W m^{-2} in the winter (January) and the minimum heat gain of 60 W m^{-2} in the spring (April). This range of values is consistent with earlier studies (Hirose, 1996; Seo, 1998). This long term net surface heat loss will be compensated by the advection of warm waters from the East China Sea.

3.3. Surface Fresh Water Flux

The surface fresh water flux is the difference between precipitation rate (P) and evaporation rate (E),

$$F = P - E. \tag{2}$$

Positive values of F indicate net water mass gain at the sea surface. The surface fresh water flux exhibits a distinct four-season pattern. The winter is characterized by fresh water gain (2 to 6 cm/mon) in the northern and northeastern JES, and fresh water loss (2 to 10 cm/mon) in the southern and southwestern JES. A strong horizontal F-gradient monotonically decreases from northeast to southwest. The spring (Fig. 5b) and summer (Fig. 5c) are both characterized by fresh water gain in the whole JES with different horizontal F-gradients: decreasing (increasing) from 4 cm/mon (4cm/mon) in the northeast JES to 2 cm/mon (6 cm/mon) in the southwest JES in the spring (summer). The autumn (Fig. 5d) is characterized by fresh water loss in the whole JES (4 to16 cm/mon) with the maximum loss of 16 cm/mon near the Tsushima/Korea Strait.

4. NUMERICAL OCEAN MODEL

4.1. Model Description

Coastal oceans and semi-enclosed seas are marked by extremely high spatial and temporal variability that challenge the existing predictive capabilities of numerical simulations. POM is a time-dependent, primitive equation circulation model rendered on a three-dimensional grid that includes realistic topography and a free surface (Blumberg and Mellor, 1987). Tidal forcing was not included in this application of the model, since high frequency variability of the circulation is not considered. River outflow is also not included. However, the seasonal variation in sea surface height, temperature, salinity, circulation, and transport are represented by the model. From a series of numerical experiments, the qualitative and quantitative effects of nonlinearity, wind-forcing, and lateral boundary transport on the JES are analyzed, yielding considerable insight into the external factors affecting the regional oceanography.

Consequently, the model contains 181 × 199 × 23 horizontally fixed grid points. The horizontal spacing is 5′ latitude and longitude (approximately 5.77 to 7.59 km in the zonal direction and 9.265 km in the meridional direction) and there are 23 sigma levels. The model domain extends from 35.0˚N to 51.0˚N, and from 127.0˚E to 142.0˚E. The bottom topography (Fig. 1) is obtained from the Naval Oceanographic Office's Digital Bathymetry Data Base 5′ × 5′ resolution (DBDB5). The horizontal friction and mixing are modeled using the Smagorinsky (1963) form with the coefficient chosen to be 0.2 for this application. The bottom stress τ_b is assumed to follow a quadratic law

$$\tau_b = \rho_0 C_D |V_b| V_b \tag{3}$$

where ρ_0(= 1025 kg/m^3) is the characteristic density of the sea water, V_b is the horizontal component of the bottom velocity, and C_D is the drag coefficient which is specified as 0.0025 (Blumberg and Mellor, 1987).

4.2. Surface Forcing Functions

The atmospheric forcing for the JES application of the POM includes mechanical and thermohaline forcing. The wind forcing is depicted by $_M$

$$\rho_0 K_M \left(\frac{\partial u}{\partial z}, \frac{\partial v}{\partial z} \right)_{z=0} = (\tau_{0x}, \tau_{0y}) \tag{4}$$

where K_M is the vertical mixing coefficient for momentum, (u, v) and (τ_{0x}, τ_{0y}) are the two components of the water velocity and wind stress vectors, respectively. The wind stress at each time step is interpolated from monthly mean climatological wind stress from COADS (1945-1989), with a resolution of 1° × 1°. The COADS wind stress was interpolated into the model grid with a resolution of 5′.

Surface thermal forcing is depicted by

$$K_H \frac{\partial \theta}{\partial z} = \alpha_1 \left(\frac{Q_H}{\rho C_p} \right) + \alpha_2 C (\theta_{OBS} - \theta) \tag{5}$$

$$K_S \frac{\partial S}{\partial z} = -\alpha_1 FS + \alpha_2 C (S_{OBS} - S) \tag{6}$$

where K_H and K_S are the vertical mixing coefficients for heat and salt, (θ, S) and (θ_{OBS}, S_{OBS}) are modeled and observed potential temperature and salinity, and c_p is the specific heat. The relaxation coefficient C is the reciprocal of the restoring time period for a unit volume of water. The parameters (α_1, α_2) are (0, 1) -type switches : $\alpha_1 1 = 1$, $\alpha_2 = 0$, would specify only flux forcing is applied; $\alpha_1 = 0$, $\alpha_2 = 1$, would specify that only restoring type forcing is applied. The relaxation coefficient C is taken to be 0.7 m/day, which is equivalent to a relaxation time of 43 days for an upper layer 30 m thick (Chu et al., 1999). The net effect is to prevent any deviation from climatology and ensure that the JES acts as a heat source.

4.3. Lateral Boundary Forcing

Boundary conditions for closed lateral boundaries, i.e., the modeled ocean bordered by land, were defined using a free-slip condition for velocity and a zero gradient condition for temperature and salinity. Thus, no advective or diffusive heat, salt or velocity fluxes occur through these boundaries.

At open boundaries, the numerical grid terminates but the fluid motion is unrestricted. Uncertainty at open boundaries makes marginal sea modeling difficult. Three approaches, local-type, inverse-type, and nested-basin/coastal-modeling, are available for determining the open boundary condition. Here, we take the local-type approach, i.e., to use the radiative boundary condition with specified volume transport. When the water flows into the model domain, temperature and salinity at the open boundary are prescribed from observational data. When water flows out of the domain, the radiation condition was applied,

$$\frac{\partial}{\partial t}(\theta, S) + U_n \frac{\partial}{\partial n}(\theta, S) = 0 \quad (7)$$

where the subscript n denotes the direction normal to the boundary.

The temperature and salinity values at the open boundaries can be either obtained from historical data such as the Navy's Master Oceanographic Observational Data Set (MOODS), or monthly mean climatological data such as the Navy's Generalized Digital Environmental Model (GDEM) data (Chu et al. 2001a, b). For simulating the seasonal variability, we use the GDEM T, S data at the open boundaries. Vertical cross-sections of monthly mean temperature (Fig. 7a) and salinity (Fig. 7b) at the Tsushima/Korea Strait show the seasonal variability of a warm-core and a salt-core. The warm-core occupies large portion of the Tsushima/Korea Strait during the winter monsoon season (November to March) and weakens and shoals during the summer monsoon season (June-August). In the summer, the water at the Tsushima/Korea Strait is strongly stratified (Fig.7a,b).

Warm water enters the JES through the Tsushima/Korea Strait with the TWC from the East China Sea and exits the JES through the Tsugaru and Soya straits. There is no evident volume transport through the Tatar Strait (Martin and Kawase, 1998), which was taken as 0 in this study. A more recent estimate of the monthly mean volume transport, reported by Yi (1996), through the Tsushima/Korea Strait with the annual average of 1.3 Sv, a maximum of 2.2 Sv in October, and a minimum of 0.3 Sv in February. The total inflow transport through Tsushima/Korean straits should be the same as the total outflow transport through the Tsugaru and Soya Straits. We assume that 75% of the total inflow transport should flow out of the JES through the Tsugaru Strait, and 25% through the Soya Strait. This ratio is adopted from the maximum volume transport through the Tsugaru Strait estimated by Toba et al. (1982), and through the Soya Strait estimated by Preller and Hogan (1998). The monthly volume transports through open boundaries are listed in Table 1.

4.4. Mode Splitting

For computational efficiency, the mode splitting technique (Blumberg and Mellor, 1987) is applied with a barotropic time step of 25 seconds, based on the Courant-Friederichs-Levy (1928) computational stability (CFL) condition and the external wave speed; and a baroclinic time step of 900 seconds, based on the CFL condition and the internal wave speed.

4.5. Two-Step Initialization

Two-steps are used to initialize the POM. During the first step (restoring run), the POM is integrated for two years from zero velocity and climatological temperature and salinity fields (Levitus 1984) with climatological monthly mean surface wind stress from the COADS data and restoring-type surface thermohaline forcing ($\alpha_1 = 0$, $\alpha_2 = 1$) which are relaxed to the surface monthly mean values. It was found that 90 days were for the model kinetic energy to reach quasi-steady state under the imposed conditions (Fig. 8). The final states of the restoring run are taken as initial conditions for the second step (simulation run). During the simulation run, the POM is integrated again for two more years with climatological monthly mean surface wind stress, net heat flux, and fresh-water flux ($\alpha_1 = 1$, $\alpha_2 = 0$) from the COADS data. The simulated temperature and salinity fields are consistent with observational studies reported in Part 1 of this paper.

5. TEMPERATURE

5.1. SST

The simulated monthly sea surface temperature (SST) is examined (Fig. 9). Comparing to the initial temperature field (Fig. 6a), the model simulated the formation of the SPF. Although SST field (Fig. 9) shows an evident seasonal variation, the SPF exists at all times throughout the year. Its position is quite stationary, but its intensity strengthens in the winter and weakens in the summer. Such a pattern is similar to earlier description (Maizuru Mar. Observ. 1997; Chu et al. 2001a, b). The location of the SPF in spring is quite consistent with Isoda and Saitoh's (1988) estimations using ten NOAA-8 satellite Advanced Very High Resolution Radiometer (AVHRR) images in spring 1984.

The SST is always found higher in the UTB than in the YB, which is consistent with Kim et al.'s (1999) observational studies. The SST gradient across the SPF is two times as strong in the winter as in the summer. The weakening of the SPF in the summer is caused by the faster warming of the water mass north of than south of the SPF in the spring. North of the SPF a second front occurs (bi-frontal structure) during the fall-to-winter transition season, especially in November and December. This front parallels the Russian coast with the maximum SST gradient around 4˚C/100 km in November.

5.2. Zonal Cross-Sections (37˚ and 43˚ N)

The zonal cross-sections (37˚ and 43˚N) of the simulated monthly mean temperature show a strong seasonal/permanent thermocline structure south of the SPF (Figure 10a) and a strong seasonal/weak permanent thermocline structure north of the SPF (Figure 10b).

South of the SPF at 37˚N (Figure 10a), the permanent thermocline is located at 80-250 m and appears all year round with the maximum strength (0.08˚C m^{-1}) in August. Above it, the seasonal thermocline occurs from the surface to 50 m depth in June (0.15˚C m^{-1}), intensifies during the summer monsoon season to a maximum value of around 0.36˚C m^{-1} in August, and weakens in September. In October, the seasonal thermocline erodes and the ocean mixed layer (OML) starts to occur. In November, the OML is well established with the temperature near 14˚C and the depth around 75 m. During the prevailing winter monsoon season (December to March), the OML deepens to 80-100 m with a westward uplift of the OML depth: 50 m near the Korean coast and 130 m near the Japanese coast. The OML starts to warm in March, and its depth shoals rapidly. The OML depth decreases from 50-100 m in March to less than 10 m in April. This process (OML warming and shoaling) continues during the summer monsoon season (June-August). The OML shoaling simulated using the POM is a month earlier comparing to the observational data (part 1 of this paper).

North of the SPF at 43˚N (Figure 10b), the permanent thermocline is quite weak. The seasonal thermocline occurs from the surface to 50 m depth in May (~0.08˚C m^{-1}), intensifies during the summer monsoon season to a maximum value of around 0.5˚C m^{-1} in August and September, and weakens in October. In November, the seasonal thermocline erodes and becomes the part of the permanent thermocline, which weakens during the prevailing winter monsoon season. In February, the permanent thermocline is so weak that the water column is almost uniformly cold (1˚C) west of 136˚E and weakly stratified (≤0.01˚C m^{-1}) east of 136˚E.

5.3. Latitudinal Cross-Section (135˚E)

The strong north-south thermal asymmetry across the SPF is also seen from the latitudinal cross-section (135˚E) of the monthly mean temperature (Figure 11). North of the SPF, the seasonal thermocline occurs near the surface in April and May, enhances drastically in summer, and is still quite strong with a vertical gradient of 0.12˚C m^{-1} in October. It weakens drastically in November. South of the SPF, the seasonal thermocline occurs in summer monsoon season, weakens in early fall, and disappears in November.

During the prevailing winter monsoon season (December to March), the simulated permanent thermocline is identified at 100 - 300 m depths south of the SPF with a vertical temperature gradient weakening from December (near 0.05˚C m^{-1}) to March (near 0.025˚C m^{-1}). The simulated permanent thermocline is much weaker north of the SPF. From January to March, there is almost no evident thermocline north of the SPF.

During the prevailing summer monsoon season (June to August), a shallow seasonal thermocline occurs with a much greater strength north of than south of the SPF; and overlays relatively uniform water north of the SPF and stratified water (the permanent thermocline) south of the SPF. North of the SPF a seasonal thermocline appears near the surface (above 50 m depth) with a vertical gradient enhancing from 0.25˚C m^{-1} in June to 0.36˚C m^{-1} in August. This strong and shallow thermocline isolates the exchange of the seawater below the thermocline from the atmospheric forcing and makes this water (north of the SPF under the thermocline) quite uniform. South of the SPF a seasonal thermocline is wider (25 - 100 m depths) and weaker with a vertical gradient around 0.13˚C m^{-1}. Such a north-south asymmetric pattern was previously presented by Kim and Kim (1999) using the Circulation Research of the East Asian Marginal Seas (CREAMS) data taken mainly in July 1995, and by Chu et al. (2001a, b) using the GDEM data.

6. Salinity

6.1. Sea Surface

The simulated monthly sea surface salinity (SSS) field (Figure 12) shows a strong seasonal variation with less (more) horizontal variability in the winter (summer). The saline Kuroshio water enters the JES through the Tsushima/Korean Strait into the JES and forms two permanent salty centers located in the northern JB (west of the Hokkaido Island) with the salinity higher than 34.0 psu, and the area between UTB and YB with the maximum salinity of 34.3 psu in August-September, respectively. The northern JB salty center has less seasonal variation than the UTB/YB salty center. Around the UTB/YB salty center, there are several fresh centers. The simulated winter (February) field (Figure 12) is consistent with that reported by Kim and Kim (1999, Figure 9 in their paper) using the data set of the Japan Oceanographic Data Center during 1930 to 1990.

6.2. Zonal Cross-Sections (37°and 43°N)

The zonal cross-sections (37° and 43°N) of the simulated monthly mean salinity show an evident sublayer (200-300 m) salinity minimum (SMIN) south of the SPF (Figure 13a) and absence of a SMIN north of the SPF (Figure 13b). This consists with many earlier studies such as Miyazaki (1952, 1953), Miyasaki and Abe (1960), Kim and Chung (1984), Senjyu (1999), Kim and Kim (1999), and Chu et al. (2001a, b).

South of the SPF at 37°N (Figure 13a), a strong seasonal halocline occurs from the surface to 30 m depth in June (0.01 psu m^{-1}), intensifies during the summer monsoon season to a maximum value of around 0.02 psu m^{-1} in August, and weakens from September to December. In January, the seasonal halocline erodes and disappears. A horizontally-oriented salinity maximum (SMAX) (S > 34.1 psu) appears above the SMIN with the interface at 200-300 m depths, and is usually broken into several salty parts with each part enclosed by the 34.1 psu isoline.

North of the SPF at 43°N (Figure 13b), the SMIN shows up in the upper layer (above 100 m) of the western JB (west of 136°E) all year round. This is consistent with Kim and Kim's (1999) identification that the East Sea Intermediate Water (ESIW) in the western JB\ is characterized by a low salinity (S < 34.00). The high salinity water (S > 34.05 psu) of the High Salinity Intermediate Water (HSIW) appears in the eastern (139°-140°E) JB. The eastern JB SMAX appears almost all year round except the spring season (March-May).

6.3. Latitudinal Cross-Section (135°E)

The strong north-south haline asymmetry across the SPF is also simulated from the latitudinal cross-section (135°E) of the monthly mean salinity: the appearance (disappearance) of the SMIN south (north) of the SPF (Figure 14).

South of the SPF, SMIN (S < 34.06 psu) occurs during the summer monsoon season

(July-October) underneath a horizontally oriented SMAX with a salty core ($S > 34.5$ psu) at 100 m depth. The SMAX with a salty core ($S = 34.3$ psu) was observed in October 1969 when a hydrographic survey for the whole JES basin was carried out by the Japan Meteorological Agency (Kim and Kim, 1999). The interface between the SMIN and the SMAX is located at 200-300 m depth. During the winter monsoon season, the SMIN is not evident.

7. CIRCULATION

7.1. General Description

The simulated surface velocity field (Fig. 15) coincides with earlier description of JES circulation presented in Section 2. The TWC separates at the Tsushima/Korea Strait into two branches through a western and an eastern channel. Flow through the western channel (i.e., EKWC) closely follows the Korean coast until it separates near 38°N into two branches. The eastern branch follows the SPF to the west coast of Japan, and the western branch, moves northward and forms a cyclonic eddy in the southern UTB. The LCC carries fresh and cold water along the Russian coast and becomes the NKCC at the North Korean coast. The NKCC meets the EKWC at about 38°N. After separation from the coast, the NKCC and the EKWC converge to a strong zonal front across the basin. A large-scale cyclonic recirculation gyre over the JB is simulated with a strong seasonal variation. It weakens and retreats northward in the spring and summer. In the fall, the cyclonic JB gyre disappears and several weak anticyclonic eddies appear. Our simulation is consistent with the earlier study (Seung and Yoon, 1995).

7.2. LCC

The LCC is a southwestward current following along the Russian coast. It bifurcates into two branches near Vladivostok: the western branch flows along the Russian-Korean coast and becomes the NKCC. The eastern branch flows southeastward, then turns eastward at 41.5˚N, and becomes the south flank of the JB gyre. The LCC has a strong seasonal variation with a maximum speed in winter and a minimum speed in the summer (Fig. 15).

Zonal cross-sections of the meridional velocity at 46˚N for January (winter), April (spring), July (summer), and October (fall) indicate seasonal and spatial variabilities of the LCC. It has a maximum southward component (0.21 m/s), occurring near the surface in winter (Fig. 16a) with a width of 100 km and the depth of 1500 m. The core of the LCC is close to the coast and near the surface. In the spring, it weakens to 0.18 m/s (Fig. 16b), but the width and depth are almost unchanged. It further weakens to a minimum of 0.08 m/s in summer (Fig. 16a) and fall (Fig. 16d). The simulated LCC in spring (Fig. 16b) is qualitatively consistent with a recent geostrophic calculation relative to the sea floor on Conductivity-Temperature-Depth (CTD) measurements (Riser et al. 1999). Moreover, there is a very weak (0.01-0.04 m/s) northward return flow on the continental shelf underneath the LCC, with a width of 30 km and the depths between 100 and 250 m.

7.3. NKCC

The NKCC is the continuation of LCC, flowing southward along the Korean coast. To clearly present its seasonal variability, we plot the simulated bi-weekly current vectors (Fig. 17) in the southwest JES (127˚- 133˚E, 35˚- 42˚N). The NKCC is evident in winter and weakens in summer.

Zonal cross-sections of the meridional velocity at 40˚N for January (winter), April (spring), July (summer), and October (fall) indicate seasonal and spatial variabilities of the NKCC. It flows along the continental slope with a maximum southward component (0.1 m/s) in winter and spring. The core of the NKCC is close to the shelf break with a width of 100 km and the depth of 1500 m (Fig. 18a). The NKCC weakens in summer (Fig. 18c) and fall (Fig. 18d).

7.4. TWC Bifurcation

The TWC enters the JES through the western and eastern channels of the Tsushima/Korea Strait. The simulated bi-weekly velocity vector fields at 10 m depth (Fig. 17) show a branching pattern. The currents through the eastern channel flow along the Japanese coast and form the FBTWC. The flow through the western channel becomes the EKWC, which closely follows the Korean coast until it separates near 37˚N into northern and eastern branches (first bifurcation). The eastern branch after the first bifurcation flows eastward until 132˚E and further separates (second bifurcation) into eastern and northern branches. The eastern branch after the second bifurcation continues to flow eastward and becomes the SBTWC. The northern branch after the second bifurcation flows northward and recirculates near 38˚N as the northern flank of a cyclonic eddy located at (130˚- 132˚E, 36˚- 38˚N).

Meanders and eddies are simulated along the four major current systems north of Korean/Tsushima Strait (EKWC, NKCC, FBTWC, and SBTWC): an evident cyclonic eddy in the southern UTB (130˚-132˚E, 36˚-38˚N) with several weak cyclonic and anticyclonic eddies north and northeast of it. The southern UTB cyclonic eddy is strong in winter with a maximum speed of 0.4m/s in January. This eddy expands northeastward from December 15 to April 15. Since then, the northwestern flank (westward flow) of this eddy weakens throughout summer. The simulated southern UTB cyclonic eddy consists with the observational study reported by (Katoh 1994; Shin et al. 1995).

7.5. Two TWC Branches Along the Japanese Coast

The TWC along the Japanese coast is characterized by strong variabilities in connection with many meanders and eddies (Fig. 15). Hase et al. (1999) identified the two main branches of the TWC along the Japanese coast using the ADCP and CTD measurements conducted east of Oki Islands every early summer of 1995-1998, and using analysis of temperature distribution at 100 m depth and the tracks of the surface drifters.

The two TWC branches along the Japanese coast are simulated in the model (Fig. 15). To present such a feature more clearly, we plotted the meridional cross-sections of the zonal velocity (Fig. 19) along 135˚E for January (winter), April (spring), July (summer), and October (fall). The FBTWC exists throughout the year. It starts from the eastern channel of the Korea/Tsushima Strait, flows along the Japanese coast, and flows out of the JES through the Tsugaru Strait (Fig. 15). Figure 19 shows that the FBTWC exists throughout the year and flows along the isobath shallower than 200 m with a maximum strength (0.3 m/s) and spatial extension (0-200 m depth, and 50 km width) in July and October, and a minimum strength (0.1 m/s) and spatial extension (0-50 m depth, and 20 km width) in April. The current flowing through the western channel of the Korea/Tsushima Strait feeds the SBTWC, which is weaker than FBTWC west of the Oki Islands (Fig. 18). The simulated SBTWC also exists throughout the year and flows along the continental shelf break with a maximum strength (0.1 m/s) in October and January, and a minimum strength (0.05 m/s) in April and July. The simulated TWC branching qualitatively coincides with recent observational studies (Hase et al. 1999). A westward flowing counter-current is also simulated below the FBTWC with the speed of 0.05 m/s (Fig. 19), which coincides with Seung's (1995) observational study.

7.6. EKWC

Zonal cross-sections of the meridional velocity at 36˚N for January (winter), April (spring), July (summer), and October (fall) indicate seasonal and spatial variabilities of the EKWC. It has a maximum northward component (0.45 m/s), occurring near the surface in fall and winter (Fig. 20) with a width of 80 km and the depth of 500 m. The core of the EKWC is close to the coast and near the surface. In the spring, it weakens to 0.3 m/s (Fig. 20), but the width and depth are almost unchanged. The simulated seasonal variability of EKWC is consistent with the observational analysis reported in the first part of this paper (Chu et al. 2001a, b).

8. CONCLUSIONS

(1) The JES circulation and thermohaline structure was simulated in this study using the POM model with the seasonal surface flux forcing. Two steps are used to initialize the POM. During the first step (restoring run), the POM is integrated for two years from zero velocity and January climatological temperature and salinity fields with climatological monthly mean surface wind stress. The final states of the restoring run are taken as initial conditions for the second step (simulation run). During the simulation run, the POM is integrated again for two more years with climatological monthly mean surface wind stress, net heat flux, and fresh-water flux from the COADS data. The simulated temperature and salinity fields are consistent with observational studies reported in Part 1 of this paper.

(2) The POM simulates the formation of the JES SPF and its seasonal variation. The simulated thermal field shows a strong north-south thermal asymmetry across the SPF. At a depth deeper than 500 m, the temperature is uniformly cold (1-2˚C). In winter, the thermocline appears in the southern JES, and disappears north of 41˚N. The strength decreases with latitude from 7˚C/100m near the Japan coast to less than 1˚C/100m at

40˚N. In the summer, a strong seasonal thermocline, caused by strong surface heating, is simulated in the shallow depths overlying the permanent thermocline. The seasonal thermocline is sustained from summer to fall.

South of the SPF, the permanent thermocline is located at 80-250 m and appears all year round with the maximum strength (0.08˚C m^{-1}) in August. Above it, the seasonal thermocline occurs from the surface to 50 m depth in June (0.15˚C m^{-1}), intensifies during the summer monsoon season to a maximum value of around 0.36˚C m^{-1} in August, and weakens in September. In October, the seasonal thermocline erodes and the ocean mixed layer starts to occur and deepens to 80-100 m with a westward uplift of the mixed layer depth: 50 m near the Korean coast and 130 m near the Japanese coast. The OML starts to warm at a rate of 2˚C/month from March to April, and its depth shoals respectively. The mixed layer shoaling simulated using the POM is a month earlier comparing to the observational data (part 1 of this paper).

North of the SPF, the permanent thermocline is quite weak. The seasonal thermocline occurs from the surface to 50 m depth in May (~0.08˚C m^{-1}), intensifies during the summer monsoon season to a maximum value of around 0.5˚C m^{-1} in August and September, and weakens in October. In November, the seasonal thermocline erodes and becomes a part of the permanent thermocline, which weakens during the prevailing winter monsoon season. In February, the permanent thermocline is so weak that the water column is almost uniformly cold (1˚C) west of 136˚E and weakly stratified (≤0.01˚C m^{-1}) east of 136˚E.

(3) The POM simulates the salinity field reasonably well with the surface featured by (a) stronger seasonal variability in the south than in the north of the SPF; and (b) two salinity activity centers located at the Tatar Strait and the Tsushima Strait.

The model also simulates a strong north-south salinity asymmetry across the SPF with the salty tongue (34.1 psu) appearance only in the south of the SPF. The shallow and strong halocline associated with strong thermocline makes the sub-surface water mass north of the SPF hydrostatically stable. The mid-level salty tongue associated with weak and wide thermocline makes the water mass hydrostatically less stable south of the SPF.

(4) The POM\ simulates the strong north-south asymmetry of the seasonal haline with an evident salinity minimum south of the SPF and absence of a salinity minimum north of the SPF. South of the SPF, the SMIN (S < 34.08 psu) occurs during the summer monsoon season (July-October) underneath a horizontally oriented SMAX with a salty core (S > 34.3 psu) at 100 m depth. The interface between the SMIN and the SMAX is located at 200-300 m depth. During the winter monsoon season, the SMIN becomes not evident. North of the SPF, the high salinity water (S > 34.05 psu) of the HSIW appears in the eastern JB (40˚-43˚N).

(5) The POM simulates the JES circulation reasonably well, especially the Tsushima Current and its bifurcation. The Tsushima Current bifurcates into a western and an eastern channel north of the Korea/Tsushima Strait. The currents through the eastern channel flow along the Japanese coast and form the first branch of TWC (FBTWC). The flow through the western channel becomes the EKWC, which closely follows the Korean coast until it separates near 37˚N into northern and eastern branches (first bifurcation). The eastern branch after the first bifurcation flows eastward until 132˚E and further separates (second bifurcation) into eastern and northern branches. The eastern branch after the second bifurcation continues to flow eastward and becomes the second branch of the TWC (SBTWC). The FBTWC exists throughout the year and flows along the isobath shallower than 200 m with a maximum strength (0.3 m/s) and spatial extension (0-200 m depth, and 50 km width) in July and October, and a minimum strength (0.1 m/s) and spatial extension (0-50 m depth, and 20 km width) in April. The simulated SBTWC also exists throughout the year and flows along the continental shelf break with a maximum strength (0.1 m/s) in October and January, and a minimum strength (0.05 m/s) in April and July.

(6) The other currents such as the LCC, EKWC, NKCC are simulated reasonably well. The simulated LCC has a maximum southward component (0.21 m/s), occurring near the surface in the winter with a width of 100 km and extending to a depth of 800 m. It weakens to a minimum of 0.15 m/s in the summer and fall, and shrinks in size to a width of 60 km and depth of 400 m. The simulated EKWC varies from 0.45 m/s (fall and winter) to 0.30 m/s (spring). The EKWC has a width of 80 km and a depth of 500 m all year round. The (northward) overshot EKWC leaves the Korean coast moving northward and converges with the southward flowing NKCC, and forms a current meandering toward east along the SPF. The simulated NKCC flows along the continental slope with a maximum southward component (0.1 m/s) in winter and spring. The core of the NKCC is close to the shelf break with a width of 100 km and the depth of 1500 m. The NKCC weakens in summer and fall.

(7) Although POM adequately simulated the JES circulation and thermohaline structure however, much more work in modeling is required. Three-dimensional observations of ocean temperature, salinity and velocity fields would allow initialization of the model with a more realistic dynamic and thermodynamic structure. Atmospheric forcing would also be more realistic by utilizing a coupled Ocean/Atmosphere model, such as the Coupled Atmosphere and Ocean Coastal System (CAOCS) under development by the Naval Postgraduate School [*Chu et al*., 1999a.] Use of such a system would provide more accurate wind stress forcing, through modifications of the wind field by surface frictional effects, inclusion of ocean wave effect, and improved ocean/atmosphere thermal and salinity fluxes.

(8) Future studies should concentrate on less simplistic scenarios. Realistic lateral transport should be included and the use of extrapolated climatological winds needs to be upgraded to incorporate synoptic winds to improve realism.

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Month	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
Soya	-0.2	-0.08	-0.08	-0.13	-0.23	-0.33	-0.43	-0.53	-0.55	-0.53	-0.48	-0.35
Tatar	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Tsugaru	-0.6	-0.22	-0.22	-0.37	-0.67	-0.97	-1.27	-1.57	-1.65	-1.57	-1.42	-1.05
Tsushima	0.8	0.3	0.3	0.5	0.9	1.3	1.7	2.1	2.2	2.1	1.9	1.4

Table 1. Monthly values of volume transport (Sv) through the lateral open boundaries. The positive/negative values mean inflow/outflow.

Figure Captions

Fig. 1. The bottom topography (m) of the Japan/East Sea (JES).

Fig. 2. Schematic map of surface current systems (after Tomczak and Godfrey, 1994).

Fig. 3. Climatological monthly mean wind stress for (a) January, (b) April, (c) July, and (d) October, using the COADS data.

Fig. 4. Climatological monthly mean net heat flux (W m^{-2}) for (a) January, (b) April, (c) July, and (d) October, using the COADS data.

Fig. 5. Climatological monthly mean precipitation minus evaporation for (a) January, (b) April, (c) July, and (d) October, using the COADS data.

Fig. 6. Initial fields from the climatological data (Levitus, 1982) at different depths: (a) temperature, and (b) salinity.

Fig. 7. Vertical cross-sections of monthly mean (a) temperature and (b) salinity at the Tsushima/Korea Strait.

Fig. 8. Temporal variation of total kinetic energy per unit volume (kg m^{-1} s^{-2}) during the model initialization.

Fig. 9. Simulated monthly mean sea surface temperature (˚C) field.

Fig. 10. Simulated zonal cross-section of the monthly mean temperature (˚C): (a) 37˚N, and (b) 43˚N.

Fig. 11. Simulated latitudinal cross-section (135˚E) of the monthly mean temperature (˚C) along 135˚E.

Fig. 12. Simulated monthly mean sea surface salinity (psu) field.

Fig. 13. Simulated zonal cross-section of the monthly mean salinity (psu): (a) 37˚N, and (b) 43˚N.

Fig. 14. Simulated latitudinal cross-section (135˚E) of the monthly mean salinity (psu) along 135˚E.

Fig. 15. Simulated monthly mean surface velocity vector field.

Fig. 16. Simulated monthly mean meridional velocity (cm s^{-1}) along 46˚N.

Fig. 17. Simulated bi-weekly surface velocity vector field in the Ulleung/Tsushima Basin.

Fig. 18. Simulated monthly mean meridional velocity (cm s^{-1}) along 40˚N.

Fig. 19. Simulated monthly mean zonal velocity (cm s^{-1}) along 135˚E.

Fig. 20. Simulated monthly mean meridional velocity (cm s^{-1}) along 36˚N.